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ORIGINAL PAPER

Temporal patterns in lacustrine stable isotopes as evidence for climate change during the late glacial in the Southern European Alps

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Abstract We investigated oxygen and carbon isotopes of bulk carbonate and of benthic freshwater ostracods (*Candona candida*) in a sediment core of Lago Piccolo di Avigliana that was previously analyzed for pollen and loss-on-ignition, in order to reconstruct environmental changes during the late glacial and early Holocene. The depth–age relationship of the sediment core was established using 14 AMS ^{14}C dates and the Laacher See Tephra. While stable isotopes of bulk carbonates may have been affected by detrital input and, therefore, only indirectly reflect climatic changes, isotopes measured on ostracod shells provide unambiguous evidence for major environmental changes. Oxygen isotope ratios of ostracod shells ($\delta^{18}O_C$) increased by ~6‰ at the onset of the Bølling (~14,650 cal BP) and were ~2‰ lower during the Younger Dryas (~12,850 to 11,650 cal BP), indicating a temporal pattern of climate changes similar to the North Atlantic region. However, in contrast to records in that region, $\delta^{18}O_C$ gradually decreased during the early Holocene, suggesting that compared to the Younger Dryas more humid conditions occurred and that the lake received gradually increasing input of ^{18}O-depleted groundwater or river water.

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Introduction

The late glacial is a period of major and rapid environmental changes that was studied intensively. In the North Atlantic region (including Central Europe) the main climatic changes include a rapid warming at the onset of the Bølling (~14,650 cal BP), a distinct cooling at the onset of the Younger Dryas (YD, ~12,850 to ~11,650 cal BP) followed by a rapid warming at the YD/Holocene transition (e.g. Rasmussen et al. 2006). The YD represents a geographically widespread climatic change and models and data agree on the involvement of the meridional overturning circulation (MOC, see for discussion Alley 2007). In the Southern European Alps, increasing oxygen-isotope ratios (δ^{18}O) from lacustrine carbonates indicate a climate warming at the onset of the Bølling but no signal could be detected from such records for the YD (Eicher 1987; Baroni et al. 2001, 2006). Still, two chironomid-inferred mean July-air temperature records indicate a warming of $\sim$3°C at the Bølling onset and a $\sim$1.5 to $\sim$2°C cooling during the YD (Heiri et al. 2007; Larocque and Finsinger 2008) and pollen records indicate vegetation responses to these climatic changes (e.g. Finsinger et al. 2006; Vescovi et al. 2007).

The Southern European Alps lie at the northern limit of the Mediterranean Basin, which is located in a climatically sensitive transitional zone between Central Europe (temperate climate) and North Africa (subtropical climate) (Fig. 1). Due to the proximity of this region to the Mediterranean Basin, Eicher (1987) suggested that the missing evidence in δ^{18}O records for the YD cooling and the subsequent climate warming at the YD/Holocene transition may be explained by a strong influence of the mediterranean climate. However, δ^{18}O records in the eastern and southern Mediterranean indicate a dry climate (marked by higher δ^{18}O values) during the YD (Bar-Matthews et al. 1997, 1999; Stevens et al. 2001; Roberts et al. 2001; Wick et al. 2003). Hence Wright et al. (2003) suggested that somewhere between Central Europe and the southern Mediterranean isotopic depletion due to cooling and isotopic enrichment due to evaporation could balance each other if cool and dry YD climatic conditions occurred.

Here we present a new record of stable isotopes measured on bulk carbonates ($\delta^{18}O_B$) and on monospecific ostracod shells ($\delta^{18}O_C$) of *Candona candida* that were collected from a well-dated sediment record of Lago Piccolo di Avigliana in the Southern European Alps. This record provides strong evidence for major late glacial and early Holocene climate variations that enable teleconnections between the southern Mediterranean and the North Atlantic climate systems.

Regional setting

Lago Piccolo di Avigliana (LPA; 45°03′ N, 07°23′ E; 350 m a.s.l.) is located in the southern foreland of the Alps (Fig. 1). The catchment (8.1 km^2) is surrounded by hills lower than $\sim$650 m a.s.l. and is characterized by metamorphic rocks (calcareous schists, serpentine, peridotite; Petrucci et al. 1970). The lake (max. depth 12.5 m, surface area 60 ha, water-residence time $\sim$0.9 years; Gaggino and Cappelletti 1984) is dammed by a moraine system deposited by a lateral

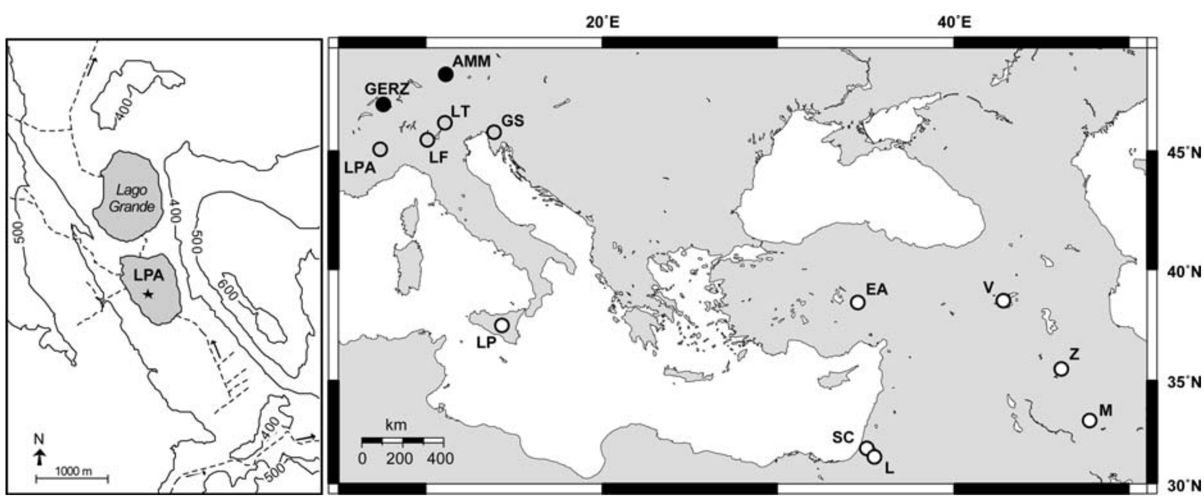

Fig. 1 *Left*: Map of the Lago Piccolo di Avigliana (LPA) area with coring location marked by the solid star. *Right*: Location of sites mentioned in the text. Southern Central Europe (filled circles): Ammersee (AMM), Gerzensee (GERZ). Southern European Alps (shaded circles): LPA, Lake Terlago (LT), Lake Frassino (LF), and Grotta Savi (GS). Mediterranean region: Lake Pergusa (LP), Eski Acıgöl (EA), Lake Van (V), Lake Zeribar (Z), Lake Mirabad (M), Soreq Cave (SC), and Lake Lisan (L)

tongue of the Susa valley glacier and receives its water from two seasonal streams and from an infilled lake (Torbiera di Trana). Mean annual temperature at Avigliana is 13.0°C, while mean temperature of the coldest and warmest months are 2.2°C (January) and 23.9°C (July), respectively (Biancotti et al. 1998). The average sum of annual precipitation amounts to 880 mm, with major rainfall occurring mainly in autumn and spring in connection with the activity of the Genoa Low, a cyclonic circulation over the Gulf of Genoa where cyclones are formed mostly in the lee of the Alps. This in turn is influenced by seasonal latitudinal shifts of the polar front (Pinna 1977). Present-day $\delta^{18}O$ of precipitation ($\delta^{18}O_P$) on the western Po Plain seems to be influenced by the shadow effect of the Alps (Dray et al. 1997) although this effect is not well defined due to a lack of collecting stations in this sector of the Alps (Longinelli and Selmo 2003). During the YD only small cirque glaciers remained in the Susa valley, none of which was located in the hydrological catchment of the lake (Carraro et al. 2002).

Materials and methods

Fieldwork and laboratory work

The lake sediments were collected with a modified piston corer (diameter: 8 cm, Merkt and Streif 1970) from a floating platform at a water depth of 12.5 m in autumn 2001 (Fig. 1b). The one-meter long drives were extruded into plastic half-tubes, wrapped in plastic foil, transported to the laboratory, stored at 4°C, and subsequently sub-sampled.

Stable oxygen and carbon isotopes were measured on bulk calcium carbonate (δ_B; 105 samples: 861–646 cm depth), adult *Candona candida* shells (δ_{CA}; 12 samples: 849–654 cm depth), and juvenile *C. candida* shells (δ_{CJ}; 17 samples: 849–654 cm depth). Sediments for δ_B were treated and measured as described by Siegenthaler and Eicher (1986). Isotopic composition of ostracod shells (δ_C) was measured with a Kiel III ThermoFinnigan device that was coupled to a MAT 250 through a custom-made communication and interface system. Sample weights ranged from 10 to 220 µg. Values are reported in the standard delta per mille notation relative to the PDB standard (‰PDB). The analytical precision is 0.1 and 0.15‰ for $\delta^{13}C$ and $\delta^{18}O$, respectively. Samples for pollen analysis (1 cm^3) were prepared by decanting and sieving at 500 µm before standard chemical treatment (HCl, KOH, HF, HCl, Acetolysis, KOH) and were eventually embedded in glycerine. Identification and counting of pollen grains (at least 400 pollen grains of terrestrial plants) was conducted under a light microscope at 400× magnification with the aid of identification keys (e.g. Moore and Webb 1991) and atlases (Reille 1992). Percentages were calculated upon the terrestrial pollen sum including tree, shrub, herb pollen, and fern spores. Loss on ignition (LOI) was measured at 550 and 950°C (following Heiri et al. 2001) to estimate the amount of organic matter, carbonate content, and minerogenic content (expressed in % of total dry weight) of the sediment. Samples for ostracod analysis were prepared as described in Belis et al. (in press).

Chronology

The depth–age relationship is based on 13 ^{14}C AMS dates on terrestrial plant macrofossils (Finsinger 2004; Finsinger et al. 2006), and one additional ^{14}C date on wood (909 cm depth; 14,930 ± 80 ^{14}C BP; Poz-6497), and by a distal micro-tephra layer, stratigraphically and chemically identified as the Laacher See Tephra (LST). The tephra was found with a discrete shard-peak concentration between 768 and 769 cm, using a modified heavy liquid-separation technique (Blockley et al. 2005). It was chemically correlated (Table 1) to the LST by WDS electron microprobe, using a Cameca SX100 microprobe (for operating conditions see Table 1). The LST has been previously reported from this region (van den Bogaard and Schmincke 1985; Schmincke et al. 1999) and is a chemically distinct phonolitic eruption that is relatively easy to distinguish from Italian eruptions of the same period (Fig. 2a). For the LST, the weighted mean age of the middle cluster (i.e. 11,066 ± 12 ^{14}C BP, Baales et al. 2002; 13,070–12,910 cal BP) was used. All radiocarbon dates were calibrated using the IntCal04 data set (Reimer et al. 2004) with the OxCal 3.1 program (Bronk Ramsey 1995, 2001). The depth–age relationship (Fig. 2b) was established by means of a generalized mixed-effect regression (Heegaard et al. 2005). The LST was given a weight equal to one in order to guarantee a correct model-estimated age. Estimated 95% confidence intervals are <200 years

Table 1 Chemical analyses of the Laacher See Tephra found at 768–769 cm depth at LPA

SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	Cl	Total
57.2	0.2	22.0	1.7	0.4	0.1	0.8	6.6	5.4	0.4	94.8
58.3	0.5	19.3	2.2	0.2	0.3	1.8	5.5	7.1	0.2	95.4
58.3	0.7	19.3	2.5	0.2	0.3	1.8	5.4	7.3	0.2	96.1
59.7	0.6	20.1	2.0	0.1	0.3	1.8	4.7	7.2	0.2	96.8
59.2	0.5	20.0	2.2	0.1	0.3	1.5	5.2	6.8	0.3	96.1
59.6	0.3	20.1	1.9	0.1	0.2	1.6	6.2	7.3	0.3	97.6
58.2	0.5	19.5	2.3	0.2	0.3	1.5	6.5	7.7	0.3	96.9
58.9	0.5	19.2	2.4	0.2	0.3	1.5	5.5	7.1	0.3	95.8
57.9	0.6	19.7	2.3	0.2	0.3	1.5	5.9	7.5	0.3	96.2
59.1	0.5	19.2	2.4	0.1	0.3	1.5	5.4	7.0	0.3	95.8
58.9	0.5	19.5	2.3	0.2	0.3	1.8	5.4	7.5	0.3	96.6
55.8	0.1	22.8	1.5	0.3	0.1	0.5	9.8	4.4	0.4	95.7
58.7	0.6	19.4	2.3	0.1	0.3	1.7	5.1	7.6	0.3	96.1
58.4	0.2	20.0	2.1	0.2	0.2	1.6	7.0	7.2	0.3	97.2
58.7	0.6	19.5	2.3	0.2	0.3	1.5	5.3	7.4	0.3	96.0
59.3	0.6	19.8	2.3	0.2	0.3	1.6	4.9	7.4	0.3	96.7

Samples were mounted in resin block and polished to a flat surface, microprobe operating conditions were 20 kev accelerating voltage, 10 na current and a 1 μm beam rastering a 10 μm spot size; Lipari Obsidian standards were used for secondary calibration and beam drift was monitored with reference to an Andradite standard. Samples with evidence of phenocryst inclusions and those with analytical totals below 95% were excluded from the results

for ages <14,000 cal BP and increase to 300 years for ages >15,700 cal BP (Fig. 2b).

Results

Changes in sediment composition

The sediment has low organic and carbonate content until 14,400 cal BP (Fig. 3h). Thereafter, carbonate and organic content increase rapidly, while the ignition residue (Fig. 3i) decreases from ca. ~90 to ~70% dry weight (% dw). Changes of smaller amplitude occur at 12,900 cal BP when organic content slightly decreased, and at 11,400 cal BP when carbonate content increased to values >20% dw.

Stable isotopes

Stable oxygen and carbon isotope ratios of bulk sediments ($\delta^{18}O_B$ and $\delta^{13}C_B$, respectively) in sediments older than ~15,500 cal BP were high (−6 and >0‰, respectively, Fig. 3c). Oxygen isotope ratios decreased to ca. −8.0‰ between ~15,200 and ~14,500 cal BP, then rapidly increased to −6.5‰ between ~14,500 and ~14,200 cal BP, and later gradually decreased to −9‰ at 9,500 cal BP. Highest $\delta^{13}C_B$ values occurred between ~14,800 and ~14,200 cal BP (~0.5‰). Between 14,600 and 14,200 cal BP $\delta^{13}C_B$ values decreased, then levelled-off at ca. −3‰ until ~11,500 cal BP, and later decreased gradually to reach −4.0‰ at 9,500 cal BP.

Pollen-inferred vegetation dynamics

With the climatic warming at the onset of the Bølling, the mixed and open woodland dominated by birch (*Betula*) and pine (*Pinus*) with large amounts of herbs (e.g. sagebrush (*Artemisia*)) and shrubs (e.g. juniper (*Juniperus*), not shown) was replaced by denser stands of *Pinus* and *Betula* (Fig. 3f–g). Summergreen oak (*Quercus*) populations expanded during the Allerød as inferred from higher pollen amounts. These oak populations partially collapsed during the YD period. Changing pollen abundances of *Artemisia*, *Quercus*, and *Betula* indicate that vegetation during the YD returned to an open woodland dominated by *Betula*, *Pinus*, and *Artemisia* and that at the onset of the Holocene oak populations recovered rapidly.

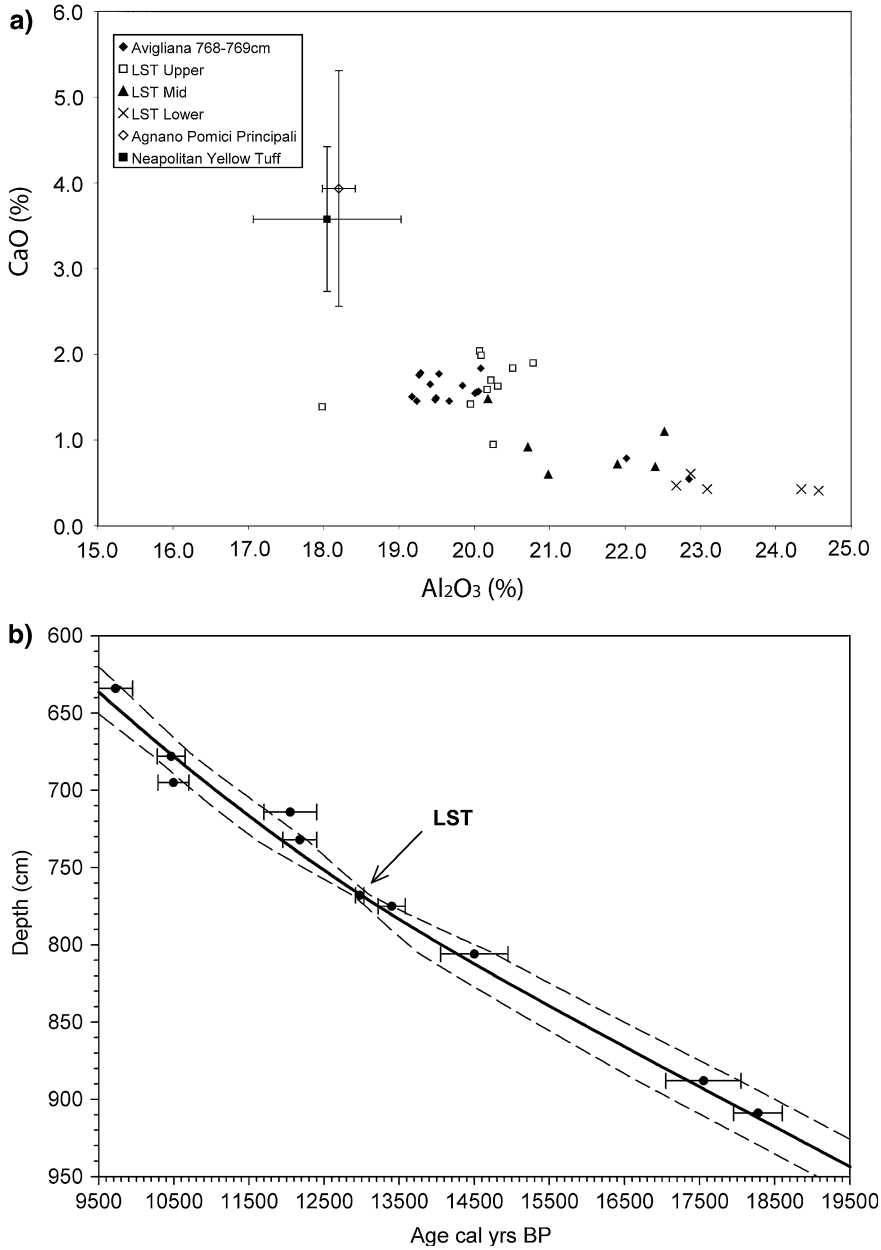

Fig. 2 (**a**) Al_2O_3 vs CaO plot for the Avigliana data compared to summary proximal data for the middle and upper phases of the Laacher See Tephra (Schmincke et al. 1999) and major Italian eruptions from the late glacial (Wulf et al. 2004), (**b**) depth–age relationship of the sediment core AVP1. The two dashed lines indicate the upper and lower 95% confidence limits of the mean expected age throughout the sequence (central line); dots and error bars: mean calibrated ages and associated 2σ errors; LST: Laacher See Tephra

Discussion

Given the high ignition residue values (Fig. 3i), it is conceivable that bulk sediment $\delta^{18}O_B$ and $\delta^{13}C_B$ may be reflecting isotopic composition of detrital carbonates. This is especially the case for samples older than 14,450 cal BP, when the sediment consisted of >80% detrital material. For this period, the $\delta^{13}C_B$ values are in the range of −2 to +1‰. Instead, for periods after 14,200 cal BP $\delta^{13}C_B$ values are around −3‰. This suggests that equilibration with atmospheric CO_2 and biological activity of plants enriched the dissolved bicarbonate with respect to groundwater (that has typically $\delta^{13}C_B$ values between −10 and −15‰, Siegenthaler and Eicher 1986; Leng and Marshall 2004) throughout the record but to varying degrees. Neglecting evaporative enrichment and $\delta^{18}O$ changes in sea water, the $\delta^{18}O_B$ composition will mainly covary

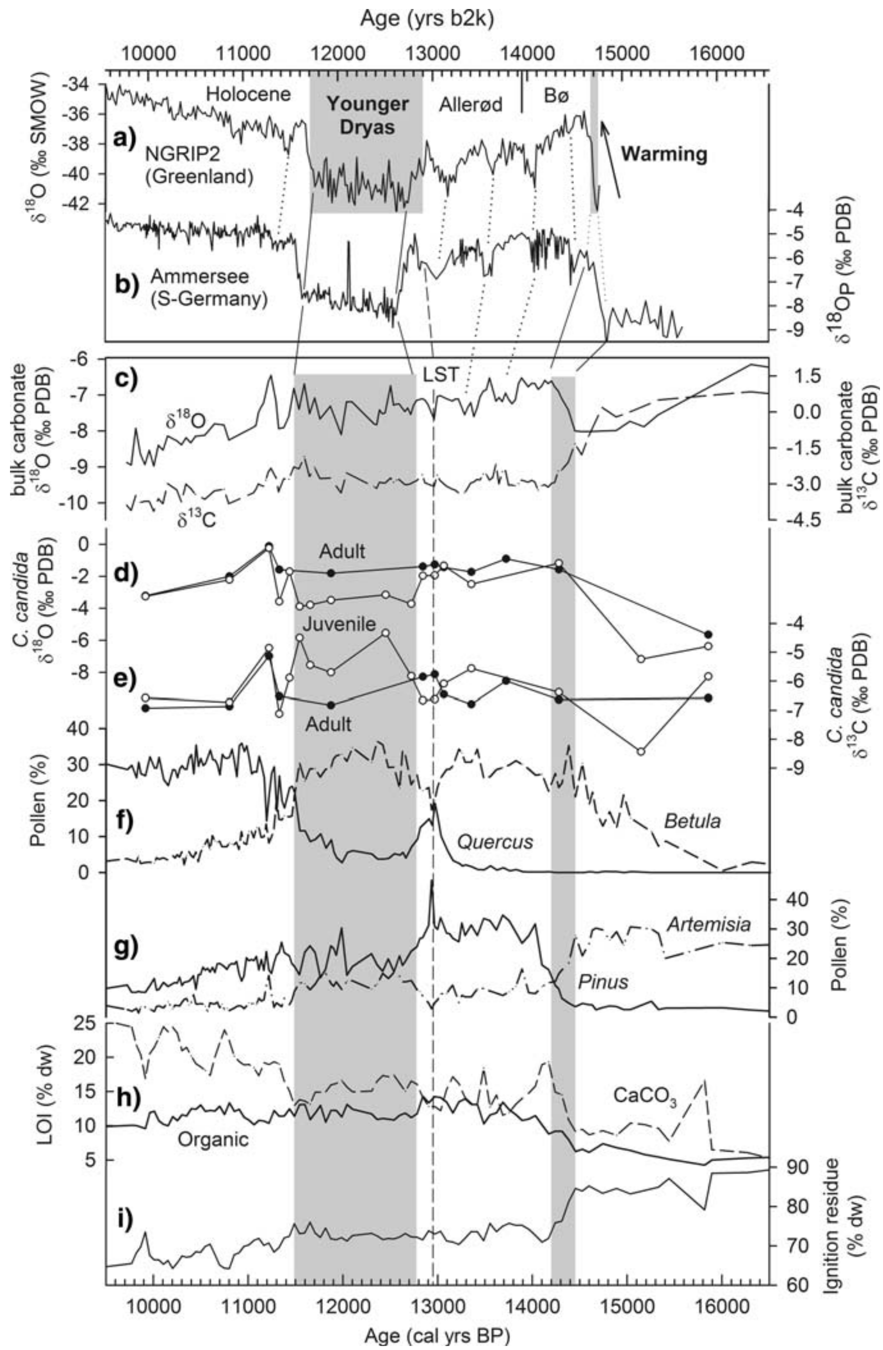

Fig. 3 Comparison between the (**a**) NGRIP2-δ^{18}O record (after NGRIP Members 2004; Rasmussen et al. 2006), (**b**) Ammersee δ^{18}O measured on benthic ostracods (von Grafenstein et al. 1999a), (**c**) stable isotopes measured on bulk carbonates (continuous line: $\delta^{18}O_B$; dotted line: $\delta^{13}C_B$), (**d**) δ^{18}O of *C. candida* shells (open circles: juveniles; closed circles: adults), (**e**) δ^{13}C of *C. candida* shells (open circles: juveniles; closed circles: adults), (**f**, **g**) selected pollen types, (**h**) organic and carbonate content (% dw), (**i**) ignition residue (% dw). LST = Laacher See Tephra. Thin grey area: onset of Bølling; large grey area Younger Dryas

with temperature—with an increase of ~0.36‰/°C (Siegenthaler and Eicher 1986). In contrast, $\delta^{13}C_B$ is indirectly related to climate due to the influence of inflowing waters (different isotopic composition), of CO_2 exchange between atmosphere and lake water, of photosynthesis/respiration of aquatic plants within the lake (Leng and Marshall 2004), and of dissolved bicarbonate from the watershed, which can be affected by changes in the relative abundance of C3 (−25‰) and C4 (−12‰) plants in the catchment (e.g. Huang et al. 2001). Variations in $\delta^{13}C_B$ ratios may parallel those of $\delta^{18}O_B$ if productivity of aquatic vegetation responded to climate change, with relatively strong (weak) biological activity in warm (cold) phases

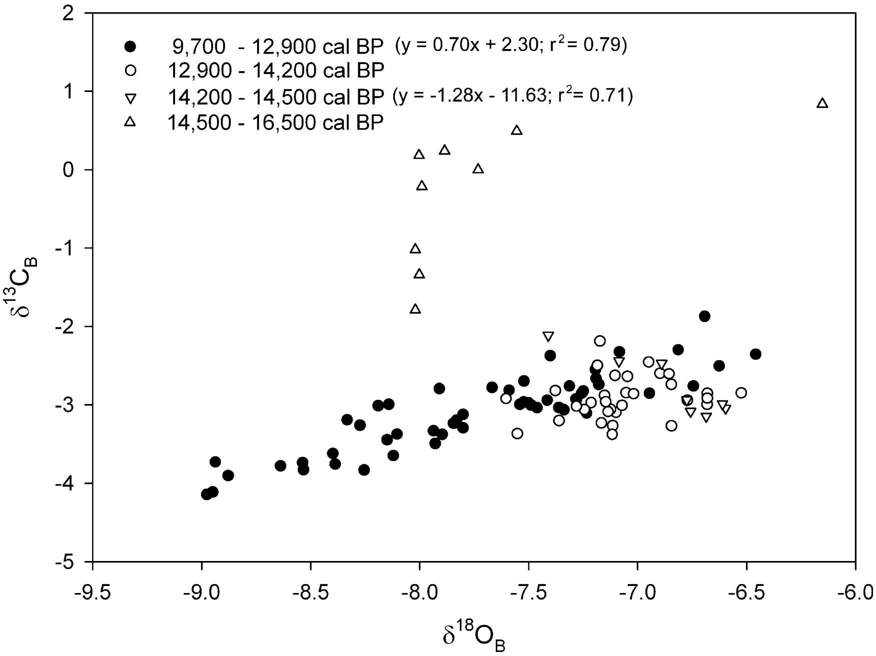

Fig. 4 Scatter plot of $\delta^{18}O_B$ vs $\delta^{13}C_B$ for Lago Piccolo di Avigliana. Samples were classified following a 10-point running correlation

leading to enhanced (reduced) ^{13}C and ^{18}O enrichment in the water (Siegenthaler and Eicher 1986). When all samples are considered the linear correlation between $\delta^{18}O_B$ and $\delta^{13}C_B$ is low (Fig. 4). However, n-point running correlations (n = 3, 5, and 10) of $\delta^{18}O_B$ and $\delta^{13}C_B$ indicate that these isotopes are negatively correlated ($r \leq -0.7$) between ~14,500 and ~14,200 cal BP. The linear correlation coefficient between ~14,200 and ~12,900 cal BP is higher ($0.02 \leq r \leq 0.6$), while $\delta^{18}O_B$ and $\delta^{13}C_B$ covary ($r \geq 0.7$) starting from ~12,900 cal BP. The decrease in $\delta^{13}C_B$ (from ~+1‰ to ~-3‰) between ~14,800 and ~14,200 cal BP may reflect the successive establishment of terrestrial vegetation (expansion of *Betula* and *Pinus*, Fig. 3f–g) and related soil development in the lake's catchment as well as the smaller influence of detrital carbonate. As discussed by Hammarlund et al. (1999) on the basis of similar carbon-isotope records from Sweden, this development can lead to increased release of ^{13}C-depleted CO_2 from soil respiration and successive depletion with time in ^{13}C of dissolved inorganic carbon (DIC) in groundwater and lakes.

Rising $\delta^{18}O_B$ values at ~14,500 cal BP may only indirectly indicate a warming climate. First, $\delta^{13}C_B$ values decreased ~200 years before the $\delta^{18}O_B$ change occurred, suggesting that the establishment of terrestrial vegetation and related soil development preceded the $\delta^{18}O_B$ change. Second, $\delta^{18}O_B$ values increase synchronously with the ignition-residue change and with the rapid closing of *Pinus–Betula* stands, suggesting that vegetation played an important role in soil stabilization thereby reducing sediment runoff. Slightly lower $\delta^{18}O_B$ values during the YD may point to a climate cooling, but the change is very subdued in this record as in other $\delta^{18}O_B$ records from the region (Eicher 1987). Decreasing $\delta^{13}C_B$ and $\delta^{18}O_B$ values starting from ~11,500 cal BP might be caused by increased input of groundwater or river water, which generally has low oxygen and carbon isotopic values. The decreasing trend was distinctly interrupted at ~11,200 cal BP by a short-term positive excursion of both isotope ratios. Since oxygen and carbon isotopes still covary during this positive excursion, the less depleted isotopic values might indicate decreased input of groundwater or river water.

In addition to bulk sediments, stable-isotope ratios of the freshwater ostracod *Candona candida* were analyzed because ostracod-isotope records are not affected by detrital carbonates. Oxygen isotope ratios of *C. candida* ($\delta^{18}O_C$) are generally less depleted (by ~4‰) than $\delta^{18}O_B$. One reason for this difference is certainly the large vital offset (ca. +2.2‰), which seems to be constant for all instars as observed by von Grafenstein et al. (1999b).

C. candida is a benthic organism (Meisch 2000), whose juveniles form their shells during the warm season while adults grow from late fall to spring. $\delta^{18}O$ values of modern juvenile Candoninae (incl. *C. candida*) shells in Ammersee (southern Germany) are depleted by up to 2‰ compared to adult shells (von Grafenstein et al. 1999b). Since the difference decreases with water depth and vanishes at about 20 m, where seasonal water-temperature differences are subdued, the difference is likely related to the temperature-dependent fractionation between calcite and water (~ -0.25‰/K, von Grafenstein et al. 1999b). Hence, if water-depth was $\sim$20 m at LPA, then the difference between juvenile and adult $\delta^{18}O_C$ would not be water-level dependent but would reflect seasonal changes in $\delta^{18}O$ of lake water. The sediments of the studied section lie at >6.5m below the present-day sediment–water interface (12.5 m water depth). Since it is unlikely that the outflow was lower than at present, we infer a maximum water depth of at least $\sim$19 m. Under this scenario, the temperature-dependent fractionation between calcite and water could be considered more or less as constant, and $\delta^{18}O_C$ would parallel the $\delta^{18}O$ of lake water, which, if the lake was an open system with a short water-residence time as it is at present, would be mainly influenced by $\delta^{18}O$ of precipitation ($\delta^{18}O_P$) (von Grafenstein et al. 1999a; Schwalb 2003). Based on these assumptions, we may infer that during most of the analyzed period the winter to summer difference remained constant since juvenile and adult $\delta^{18}O_C$ values are similar. A summer-$\delta^{18}O_P$ decrease during YD (12,850–11,650 cal BP) can be inferred from lower $\delta^{18}O_{CJ}$, while little can be said for winter (only one sample). Increasing $\delta^{13}C_{CJ}$ during this 1200-year-long cold phase might have been caused by a reduction of the summer inflow (hence drier climate), which would lead to an increase of $\delta^{13}C_{DIC}$ in the lake. This is because DIC of inflowing water, with relatively negative $\delta^{13}C$, compensates both lake-internal processes of enrichment (net-sedimentary flux of organic matter and equilibration with atmospheric CO_2, von Grafenstein et al. 2000). A decreased summer-water input might also be inferred from the abundance decrease of *C. candida* and the occurrence of *Darwinula stevensoni* (Belis et al. in press), which at present is an indicator of shallow water (0–12 m water depth) in modern lakes (Meisch 2000). After $\sim$11,500 cal BP, $\delta^{18}O_C$ gradually decrease, only being interrupted by a positive shift at $\sim$11,200 cal BP. During the latter event, higher $\delta^{18}O_C$ and $\delta^{13}C_C$ might indicate a return to pre-YD $\delta^{18}O_P$ conditions. However, since similar changes are depicted in δ_B ratios, decreasing δ_C values during the early Holocene may as well have been caused by increasing input of groundwater or river water, or by a concomitant onset of eutrophication in the lake.

Major changes in vegetation composition at LPA, as inferred from the pollen record, are contemporaneous with changes in $\delta^{18}O_C$ and $\delta^{13}C_C$ ratios (within the resolution limits of the $\delta^{18}O_C$ record; Fig. 3). At the onset of the Bølling, $\delta^{18}O_C$ increased when mixed-woodland dominated by *Betula*, *Pinus*, *Larix*, and *Juniperus* were replaced by denser stands of *Pinus* and *Betula*. During the YD $\delta^{18}O_C$ decreased along with a partial collapse of mixed forests (e.g. *Quercus*) and the expansion of *Betula* and *Artemisia*-dominated shrubland.

An increase at the onset of the Bølling, followed by a decrease during the YD and a subsequent increase at the YD/Holocene transition (YD/PB) has been observed in several $\delta^{18}O$ records of lacustrine carbonates in Central and northern Europe (e.g. Eicher and Siegenthaler 1976; Lotter et al. 1992; von Grafenstein et al. 1999a) and of ice-cores in Greenland (e.g. NGRIP, Rasmussen et al. 2006). The change in $\delta^{18}O$ values is interpreted as reflecting changes in air temperatures. For the lacustrine records the main assumption states that variations of the drainage-basin water balance (precipitation−evaporation) and of the long-term humidity have only negligible effect (Eicher and Siegenthaler 1976; von Grafenstein et al. 1999a). A similar sequence of July-air temperature changes is confirmed for the Southern European Alps by chironomid-inferred temperature records at LPA and Lago di Lavarone (Heiri et al. 2007; Larocque and Finsinger 2008).

In contrast to Central European $\delta^{18}O$ records, an increase in the $\delta^{18}O$ values during the YD and a strong decrease in the $\delta^{18}O$ values at the YD/PB transition has been observed in several records from the Mediterranean region (Fig. 1), like in the speleothem record of the Soreq Cave and in lacustrine carbonates in Lake Lisan, Israel (Bar-Matthews et al. 2003; Kolodny et al. 2005), in Lake Van and in Eski

Acıgöl, Turkey (Roberts et al. 2001; Wick et al. 2003), in Lake Zeribar and Lake Mirabad, Iran (Stevens et al. 2001, 2006), in Lake Pergusa, southern Italy (Zanchetta et al. 2007), and in the speleothem record from Grotta Savi, northeastern Italy (Fisia et al. 2005). A decrease in $\delta^{18}O$ ratios during the early Holocene has also been observed in two records from the southeastern Alps: at Lake Terlago (Baroni et al. 2001) and at Lake Frassino (Baroni et al. 2006). These records have been interpreted as reflecting changes in the seasonality of precipitation (Lake Zeribar and Lake Mirabad), changes in the isotopic composition of source moisture and rainfall amount (Soreq Cave, Lake Lisan, and Lake Pergusa), or changes in the water balance of the lake (Eski Acıgöl), the amount of rainfall (Lake Frassino and Lake Terlago), and the combined effect of temperature and rainfall (Grotta Savi).

During the YD, as during other North Atlantic cold spells (e.g. Heinrich events), climatic conditions were generally drier in the southern Mediterranean region (e.g. Lamb et al. 1995; Combourieu Nebout et al. 2002) leading to higher $\delta^{18}O$ ratios in lacustrine and speleothem carbonates. Wright et al. (2003) suggested that somewhere between the southern Mediterranean and Central Europe isotopic depletion due to temperature decrease and isotopic enrichment due to evaporation could have balanced each other if cool and dry YD climatic conditions occurred. As suggested by our $\delta^{18}O_C$ record, which shows more negative values during the YD, this is clearly not the case for the Southern European Alps although chironomid-inferred temperatures indicate a temperature decrease (Heiri et al. 2007; Larocque and Finsinger 2008) and higher $\delta^{13}C_C$ values suggest a decreased inflow.

It is noteworthy that, within the limits of the stratigraphic resolution, the LPA $\delta^{18}O_C$-inferred climate changes during the late glacial show similar changes as those reconstructed for the same time-slice in Central Europe (e.g. Ammersee, von Grafenstein et al. 1999a, Fig. 4) and in the North Atlantic (Alley 2000). In contrast, soon after the onset of the Holocene, the patterns at LPA and Central Europe diverge with decreasing $\delta^{18}O_C$ at LPA and increasing $\delta^{18}O$ ratios north of the Alps. Thus, $\delta^{18}O_C$ changes at LPA do not reflect a typical 'Central European' (e.g. Ammersee, Fig. 4) or 'Mediterranean' pattern (e.g. Soreq Cave). Instead, they seem to reflect a 'Central European' pattern for the late glacial (increasing values at the onset of the Bølling and decreasing $\delta^{18}O_C$ values during the YD) and a 'Mediterranean' pattern for the early Holocene (decreasing $\delta^{18}O_C$ values).

Conclusions

The Lago Piccolo di Avigliana sediments reflect the impact of major late glacial and early Holocene climate changes on sediment and vegetation composition, stable isotopes of bulk sediments, and more clearly, on stable isotopes of monospecific freshwater ostracod (*C. candida*) shells. While oxygen and carbon isotope ratios of bulk sediments may be affected by detrital carbonates and, therefore, only indirectly record climate changes, stable isotopes measured on ostracods provide evidence for the onset of the Bølling warming and the YD cooling. During the late glacial the climate changes were, therefore, similar to those documented in Central Europe, as also inferred using chironomid-based temperature reconstructions in the Southern European Alps. In contrast, during the early Holocene the record does not show a climate warming as other records do in Central Europe, but rather suggests that the lake received gradually increasing input of groundwater or river water. This is also reflected in other stable isotope records from the Southern European Alps and from the Mediterranean region. However, many questions remain concerning climate changes and their impact on terrestrial and aquatic environments of the region. Further efforts are required to develop higher-resolution climate reconstructions in order to highlight late glacial and Holocene centennial-scale climate changes and changes in climate gradients across Southern and Central Europe.

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